Modification of the Baroclinic Instability Associated with Positive and Negative Arctic Oscillation Index: A Theoretical Proof of the Positive Feedback

doi: 10.2151/sola.2011-014

Modification of the Baroclinic Instability Associated with Positive and Negative Arctic Oscillation Index: A Theoretical Proof of the Positive Feedback

Sawaka Seki[1], H. L. Tanaka[2], and Fuyuki Fujiwara[1]

[1]*Graduate School of Life and Environmental Sciences*, *University of Tsukuba*, *Tsukuba*, *Japan*

[2]*Center for Computational Sciences*, *University of Tsukuba*, *Tsukuba*, *Japan*

Abstract

The modification of the baroclinic instability associated with positive and negative Arctic Oscillation Index (AOI) is theoretically investigated using a linearized 3D spectral primitive equation model. The linear instability analysis shows that the most unstable Charney mode M_C changes its structure to intensify (weaken) the polar jet by the eddy momentum flux associated with the positive (negative) AOI. More importantly, the meridionally dipole Charney mode M_2 is modified into the monopole Charney mode M_1 to transport eddy momentum flux northward under the positive AOI condition. It is found that this modification is essential to intensify the polar jet during the AOI positive phase. Hence, we have theoretically confirmed that there are positive feedbacks between the baroclinic instability waves and the Arctic Oscillation characterized by the intensity of the polar jet.

1. Introduction

The Arctic Oscillation (AO) is a north-south seesaw pattern of the atmospheric mass between the arctic region pole-ward of 60°N and a surrounding zonal area in mid-latitude (Thompson and Wallace 1998). The AO is defined as a primary mode of empirical orthogonal function (EOF-1) for the sea level pressure (SLP) fields in the Northern Hemisphere (NH). The AO Index (AOI) is then defined as the time series for EOF-1 scores associated with the SLP. The zonal wind pattern of the AO is characterized by the strength of a subtropical jet and a polar jet. When the AOI is positive, the polar jet tends to be strong and the subtropical jet tends to be weak, and vice versa.

Many previous studies indicated that there are positive feedbacks between the AO and the transient baroclinic waves. For example, Yamazaki and Shinya (1999) found that the wave forcing contributes to the transition to the high/low polarity of the AO. This means that the zonal-eddy interaction plays an important role for the transition of the AO. They also showed that the zonal wave components of planetary-scale wavenumber 2 and 3 contribute most to the transition. Lorenz and Hartmann (2002) observationally investigated the variability of the zonal-mean zonal wind in the NH winter using EOF analysis and momentum budget diagnostics. They concluded that the baroclinic eddies are most important for the positive eddy-zonal flow feedback and the quasi-stationary eddies also reinforce the zonal wind anomalies. Moreover, Limpasuvan and Hartmann (1999) demonstrated the important role of transient and stationary eddy fluxes in the maintenance of annular modes of variability with a realistic numerical simulation. They found that the eddy structures evolve with the jets and the transient eddy momentum flux supports the shifts in jet position as the most important eddy momentum forcing. In their study, the ridge or trough axes tend to tilt from southwest to northeast when the jet shifts northward in the North Atlantic. Then the associated stationary wave transports momentum pole-ward to support the local jet displacement. These studies argue that the positive feedbacks between the zonal-mean wind anomalies and the baroclinic eddies are important for the AO. However, many of these studies are classified into the observational studies rather than theoretical studies.

On the other hand, Tanaka and Tokinaga (2002) theoretically investigated the development of the baroclinic instability with the variable basic states depending on the AOI using a 3D spectral primitive equation model in terms of the 3D normal mode expansion. They found that the baroclinicity of the strong polar jet, which is associated with the positive AOI, excites a monopole Charney mode (M_1) centered at high-latitudes. M_1 is dynamically the same ordinary Charney-type baroclinic instability as M_C (Charney 1947) but is excited by the baroclinicity of the polar jet. The structure of M_1 shows the transport of the westerly momentum to the polar jet. Since M_1 was replaced with the other most unstable mode only for the strong polar jet, it was theoretically explained that there is a positive feedback between M_1 and the polar vortex. However, they investigated only the development of the baroclinic waves under the context of extreme events for the past AOI.

In this study, we investigate the modification of the baroclinic instability associated with the positive or negative AOI. The basic method is following Tanaka and Tokinaga (2002). The baroclinic instability problem is solved for the AO-related basic states. Finally we theoretically explain the positive feedback between the baroclinic instability and the AO. In Section 2 the data and governing equations are described for the linear baroclinic instability problem. The results of the stability analysis are presented for the most baroclinically unstable mode in Section 3. Ultimately the conclusion is summarized with discussion in Section 4.

2. Data and method

2.1 Data for the zonal-mean basic state

The data used in this study are National Centers for Environmental Prediction (NCEP)-National Center for Atmospheric Research (NCAR) reanalysis of four-times-daily zonal wind at constant pressure levels for December to February (DJF) during 1958−1997 (Kalney et al. 1996). Figure 1a shows the climatology of zonal-mean zonal wind in January during this period. In order to construct the basic states representing the AOI positive and negative, we added (or subtracted) the zonal wind anomaly regressed with the normalized AOI (Fig. 1b) to the climatology of the zonal-mean zonal wind (Fig. 1a). For example, adding the three-fold zonal wind anomaly to the climatology, we constructed the basic state of zonal-mean zonal wind for the AOI +3.0σ (standard deviation) in Fig. 1c. Although Fig. 1c is an artificial basic state for the AOI +3.0σ, it is almost identical to the observation for DJF in 1988/89 (see Tanaka and Tokinaga 2002). Similarly, the artificial zonal-mean zonal wind for the AOI −3.0σ (Fig. 1d) represents the realistic one for the large negative AOI. In this study, the basic states of the AOI ranging from −3.0σ to +3.0σ are prepared for the linear instability analysis.

Corresponding author: Hiroshi L. Tanaka, Center for Computational Sciences, University of Tsukuba, 1-1-1 Tennoudai, Tsukuba, Ibaraki 305-8577, Japan. E-mail: tanaka@sakura.cc.tsukuba.ac.jp.

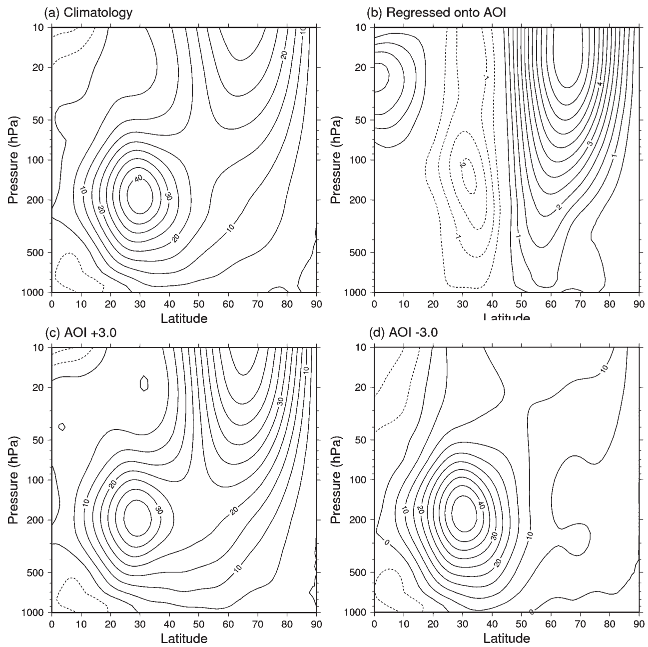

Fig. 1. Latitude-height structure of (a) zonal wind climatology (m s^{-1}) for January 1958 to 1997 in NH, (b) zonal wind anomaly (m s^{-1}) regressed onto the AOI index for DJF in 1958 to 1997, (c) the zonal-mean basic state of zonal wind (m s^{-1}) for the AOI +3.0σ, and (d) as in (c) but for the AOI −3.0σ. Solid line means positive value and dotted line means negative value.

2.2 Model

In order to amplify the baroclinically unstable wave for the basic states associated with the positive and negative AOI and to solve the linear instability analysis, we adopted a 3D spectral primitive equation model in terms of the 3D normal mode expansion. The detail of the model description was provided by Tanaka and Kung (1989). Since we followed the procedure demonstrated by Tanaka and Tokinaga (2002), we present a brief description here.

Transforming the primitive equations on a sphere into the wavenumber space with 3D normal mode functions, and linearizing them with respect to a given zonal basic state, we obtain a prognostic equation for the anomaly of the normal mode expansion coefficients w_i as:

$$\frac{dw_i}{d\tau} = -i\sigma_i w_i - i\sum_{j=1}^{K}\left[\sum_{k=1}^{K}(r_{ijk}+r_{ikj})\overline{w}_k\right]w_j, \quad i = 1, 2, 3, \ldots, K, \tag{1}$$

where τ is a dimensionless time, symbol i is an imaginary unit, σ_i is the eigenfrequency of Laplace's tidal equation, r_{ijk} are the interaction coefficients for nonlinear wave-wave interactions, and subscripts i (or j, k) denote the 3 dimensional wavenumbers. The external forcing for transient components is disregarded in this study. Although the model is based on the primitive equations on a sphere, we can filter out the redundant unstable gravity modes due to the normal mode expansion technique.

The AO-modified basic states are substituted into the zonal basic state ($\overline{w}_k$) as a prescribed parameter. Since σ_i and r_{ijk} are determined accurately from the primitive equations, we can solve the linear equation as an eigenvalue problem by assuming a wave-type solution for the prognostic equation:

$$w_i(\tau) = \psi \exp(-i\nu\tau), \tag{2}$$

where ψ is a complex-valued eigen vector describing the structure of the unstable mode, and ν is an eigenvalue whose real and imaginary parts represent the phase speed and the growth rate for the unstable modes, respectively. The matrix to be solved becomes a block diagonal for each zonal wavenumber so that we can solve the eigenvalue problem for each zonal wavenumber. We used vertical modes of m = 0 to 6 and meridional modes l = 0 to 20 but for the symmetric Rossby modes. The problem of the baroclinic instability can thus be solved for arbitrary zonal basic states on a sphere by this method in the framework of the primitive equation. The fastest growing mode represents the most unstable mode.

The structures of the unstable modes may be analyzed by the inverse Fourier transform of ψ from the wavenumber space to the physical space using the 3D normal mode functions. Selecting the most unstable modes, we can analyze the relationship between the positive and negative AOI and the baroclinically unstable waves.

3. Results

First, we demonstrate the baroclinic instability for each artificial basic state. Figure 2 illustrates the growth rates (day^{-1}) and phase speeds (° day^{-1}) of unstable modes computed for the zonal-mean basic state of the AOI +3.0σ (Fig. 1c). In the synoptic scale of zonal wavenumber n = 6−12, the fastest growing unstable mode is Charney type monopole mode M_C (see Charney 1947; Tanaka and Kung 1989). However the most unstable mode in the planetary scale of n = 1−5 is replaced by the other monopole mode M_1 which is located at higher latitude than M_C (see Tanaka and Tokinaga 2002). Meridional dipole Charney mode M_2 also appears in the planetary scale (see Tanaka and Kung 1989). We find tripole Charney mode M_3 in other basic states (not shown). M_2 and M_3 are identified as the high-order meridional modes of M_C that have nodes in the meridional direction. These characteristics described in Fig. 2 agree with Fig. 9 in Tanaka and Tokinaga (2002) showing the growth rates and phase speeds of each unstable mode related to the observed basic state for the AOI +3.0σ. This agreement is also seen in the cases of other basis states.

In order to find the relationships between the AOI and the baroclinic instability waves, we analyzed the most unstable modes at each scale. In this paper, M_C at n = 6 and M_2 at n = 2 are discussed in detail since the characteristics of their modal structure are represented by these two examples.

3.1 Charney mode M_C

Figure 3a illustrates the 500 hPa geopotential deviations from the zonal mean in the NH for M_C at zonal wavenumber 6 with the climatological basic state (Fig. 1a). The maxima of geopotential amplitudes are located at the mid-latitudes. In the northern part of the maxima their ridge (or trough) axes tilt from northwest to southeast and in the southern part their axes tilt from northeast to southwest. This means that westerly eddy momentum is transported and converged at the mid-latitudes. The momentum flux thus intensifies the subtropical jet.

On the other hand, Fig. 3b shows the same structure but for the basic state of AOI +3.0σ (Fig. 1c). The growth rate is approximately 0.27 day^{-1} (Fig. 2a). It moves eastward with its phase speed of approximately 8.6 ° day^{-1} (Fig. 2b). The ridge axes predominantly tilt from northeast to southwest. Although the amplitude maxima stay at mid-latitudes even if the AOI becomes large positive (Fig. 3b), the tilted axes tend to transport westerly momentum northward and reinforce the polar jet. Hence, the structure of M_C is modified to intensify the polar jet by the enhanced northward eddy momentum flux so that the AOI becomes positive. M_C in other zonal wavenumbers also indicate the same modification.

To investigate whether the modification in the modal structure is continuous or not, the growth rates for each mode are examined as a function of the AOI (Fig. 3c). According to the result, the growth rate of M_C continuously changes with varying in the AOI. The same characteristics are found in the phase speeds of M_C (not shown). M_C continuously changes its structure from a typical Charney mode to a different shaped Charney mode transporting eddy momentum poleward.

Therefore, as the AOI becomes positive, M_C transports more

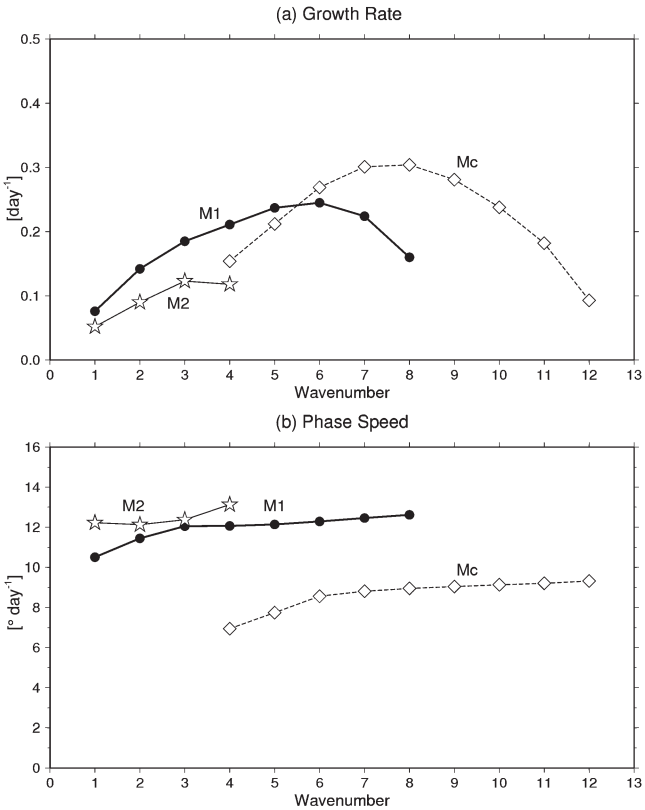

Fig. 2. (a) Growth rates and (b) phase speeds of the first-three unstable modes (labeled as M_C, M_1, and M_2) as functions of zonal wavenumber 1−13 for a zonal basic state of the AOI +3.0σ.

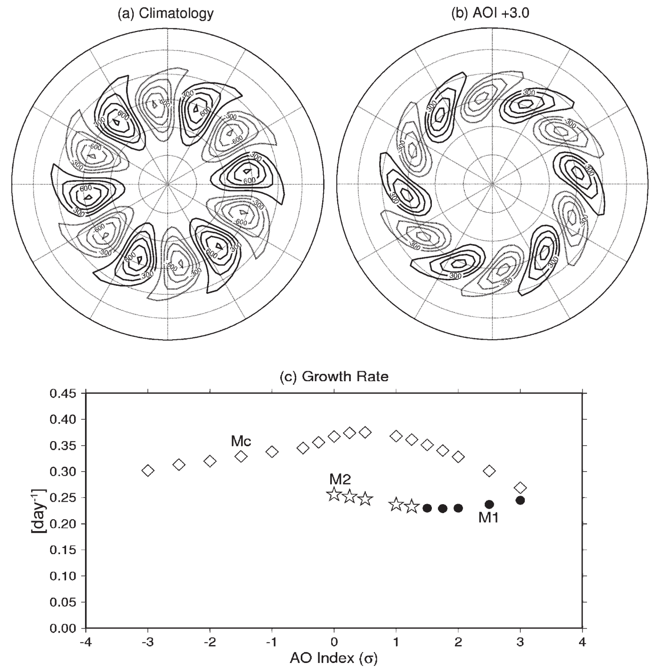

Fig. 3. 500 hPa geopotential deviation (in arbitrary unit) from the zonal mean for M_C in zonal wavenumber 6 for (a) climatological basic state and (b) AOI +3.0σ in the NH. The solid lines are positive deviations and the bold dotted lines are negative deviations. (c) The growth rates of the unstable modes in zonal wavenumber 6 as a function of the AOI for three types of unstable modes.

westerly momentum northward to intensify the polar jet inducing the positive AOI. This theoretically shows that there is a positive feedback between the AO and baroclinically unstable Charney mode M_C. In addition, M_2 is also modified into another monopole mode as the AOI becomes positive in Fig. 3c. We will refer to this in the next section.

3.2 Dipole Charney mode M_2

Figure 4 illustrates latitude-height structures of geopotential amplitudes and their phase (longitudes of ridges) for the dipole mode M_2 at zonal wavenumber 2. When the AOI is negative with a weak polar vortex, it is obvious that M_2 has the meridional dipole structure with two amplitude maxima in the north and south (Fig. 4a). But the amplitude of the southern maximum gets weaker as the AOI increases to positive (Figs. 4b, 4c). And eventually M_2 is modified to the monopole structure with only northern pole (Fig. 4d). Compared with M_C for the high AOI (Fig. 3b), the northern amplitude maximum of M_2 is located at very high latitudes (Fig. 4d). Its ridge axes also tilt in the horizontal plane to feed the eddy momentum flux northward as in the case of M_C. Similarly, the phase tends to tilt westward with respect to height at the high-latitudes as the AOI becomes positive (Figs. 4c, 4d). Since the wave activity flux is perpendicular to the phase contour in the figure, we can infer the upward and southward wave-activity flux for Fig. 4d. It is found that the eddy momentum flux extends to higher latitudes as the AOI increases. The modified monopole structure is identical to the monopole Charney mode M_1 advocated by Tanaka and Tokinaga (2002). It is found in this study that M_1 is not a distinct mode associated with the baroclinicity of the polar jet, but is identified as M_2 modified to a monopole structure by the positive AOI.

Consequently, M_2 is modified into a monopole structure or M_1 to intensify the polar jet as the AOI becomes large positive. This implies that the unstable planetary mode M_2 shows a positive feedback with the AOI.

Figure 5 illustrates the growth rates for zonal wavenumber 2 as a function of the AOI for various types of unstable modes. The growth rates show that M_2 is modified to M_1 continuously as the AOI becomes large positive. In addition, M_3 is modified to M_2 as the AOI becomes large positive. Of course, these features also appear in the variation of the phase speed (not shown). The same modification in the structure described for M_2 at the zonal wavenumber 2 is also seen for other zonal wavenumbers.

4. Conclusion and discussion

The AO is characterized by the intensity of the polar jet and polar vortex. Observational studies revealed that the westerly jet is built up by the meridional flux convergence of the eddy momentum provided by the baroclinic instability waves (Limpasuvan and Hartmann 1999; Yamazaki and Shinya 1999; Lorentz and Hartmann 2002). In this study, the relationship between the AO and baroclinic instability waves is examined theoretically using a 3D spectral primitive equations linearized with respect to the zonal basic states for various positive and negative AOI.

As a result, the basic state with a positive AOI modifies the ridge axes of M_C and M_2 to tilt from northeast to southwest. Thus the eddy momentum is transported further to the north to enhance the polar jet. Moreover, the southern amplitude maximum of M_2 disappears as the AOI becomes positive. The single amplitude maximum is located at higher latitudes than those of M_C so that M_2 becomes a monopole structure or the monopole Charney mode M_1 documented by Tanaka and Tokinaga (2002). M_1 plays an important role to produce a persistently strong polar jet. Therefore, it is theoretically shown that there are positive feedbacks between the AO and baroclinically unstable waves. We also find in this study that M_1 is not a distinct mode associated with the baroclinicity of the polar jet, but is identified as M_2 modified to a monopole structure by the positive AOI. The different structures in the eddy

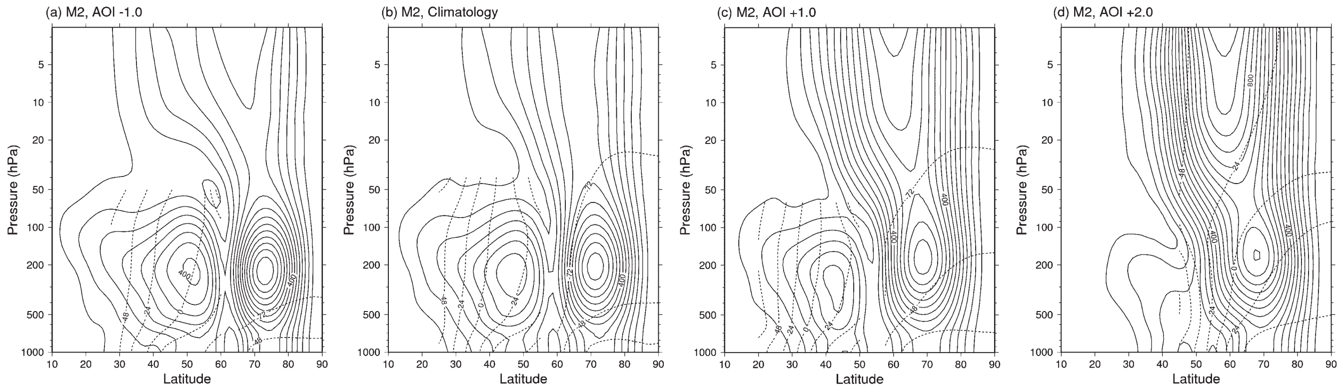

Fig. 4. Latitude-height structures of geopotential amplitudes (solid line in arbitrary unit) and phases (dotted line denoting longitudes of ridges) for M_2 at zonal wavenumber 2 for (a) AOI -1.0σ, (b) climatological basic state, (c) AOI $+1.0\sigma$, and (d) AOI $+2.0\sigma$ in NH.

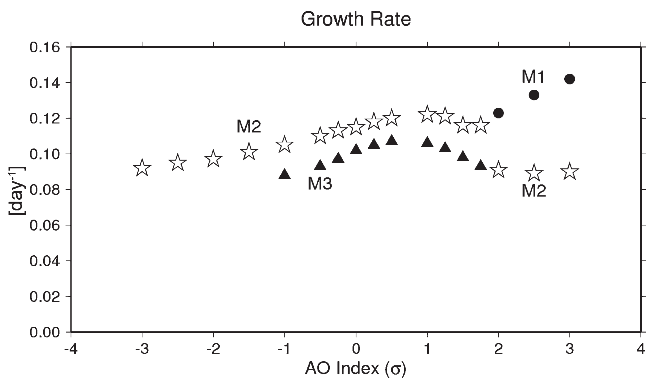

Fig. 5. Growth rates for the various types of the unstable modes at zonal wavenumber 2 as a function of the AOI.

momentum flux for the positive and negative AOI are presented observationally by Ikeda and Tanaka (2008), which support quantitatively with the present theoretical explanation.

The positive feedbacks accelerate the polarity of the AO. However, the extremely large value of the AOI must be terminated by the nonlinearity of the atmosphere, which is beyond the scope of the present linear theory with the 3D spectral primitive equation model. Also the wave-wave interactions for the non-zonal basic state may be important to modify the structure and behavior of the baroclinic waves. As a future subject, we need to expand the problem for the general non-zonal basic state to construct a general linear baroclinic model.

Acknowledgement

All figures are generated using Generic Mapping Tool. The 3D spectral primitive equation model Eqs. (1) and (2) are solved with International Mathematics and Statistics Library.

References

Charney, J. G., 1947: The dynamics of long waves in a baroclinic westerly current. *J. Mereor.*, **4**, 135−162.

Ikeda, M., and H. L. Tanaka, 2008: Positive feedback between polar jet and polar mode of the baroclinic instability. *Proc. The First Int'l Symposium on the Arctic Research.*, Nov. 2008, Tokyo Japan, 210−213.

Kalnay, E., and co-authors, 1966: The NCEP/NCAR 40-Year Re-analysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437−471.

Limpasuvan, V., and D. L. Hartmann, 1999: Eddies and the annular modes of climate variability. *Geophys. Res. Lett.*, **26**, 3133−3136.

Lorenz, D. J., and D. L. Hartmann, 2003: Eddy-zonal flow feedback in the northern hemisphere winter. *J. Climate*, **16**, 1212−1227.

Tanaka, H. L., and E. C. Kung, 1989: A study of low-frequency unstable planetary waves in realistic zonal and zonally varying basic states. *Tellus*, **41A**, 179−199.

Tanaka, H. L., and H. Tokinaga, 2002: Baroclinic instability in high latitudes induced by polar vortex: a connection to the arctic oscillation. *J. Atmos. Sci.*, **59**, 69−82.

Thompson, D. W. J., and J. M. Wallace, 1998: The arctic oscillation signature in the wintertime geopotential height and temperature fields. *Geophy. Res. Lett.*, **25**, 1297−1300.

Yamazaki, K., and Y. Shinya, 1999: Analysis of the arctic oscillation simulated by AGCM. *J. Meteor. Soc. Japan*, **77**, 1287−1298.

Manuscript received 1 February 2011, accepted 21 March 2011
SOLA: http://www.jstage.jst.go.jp/browse/sola